INTEGRATED STRUCTURAL SUCCESSION AND AGE CONSTRAINTS ON A SVECOFENNIAN KEY OUTCROP IN VÄSTERVIKEN, SOUTHERN FINLAND

by
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The Västerviken outcrop in Karjalohja, southwestern Finland, represents an example of polyphase migmatitic rocks of the Svecofennian amphibolite facies supracrustal gneiss terrain, southeast of the non-migmatitic rocks in the Orijärvi area. The structural sequence on the outcrop yields two major groups of ductile deformation events, Eₐ-E₋ and E₄-E₅, with a cooling stage in between them and later wide-scale open warping structures, Dₐ-D₉, of the earlier structures. The structural succession at Västerviken is correlated to the regional structural scheme in the southern Svecofennian domain.

The primary features of the felsic gneiss suggest a volcanic or volcano-elastic origin at c. 1.9 Ga. The earliest structures identified are isoclinal, intrafolial folds, Fₐ, in layered felsic gneisses, but they were not identified in the D₉-deformed gneissic tonalities. The first strong metamorphic pulse was accompanied by structures that provide evidence of an intense shortening event, Eₐ, that caused isoclinal Fₐ folding of the polyphase neosome veins at c. 1.88 Ga. Discordant relations of mafic dykes to the host rock and to Eₐ neosome veins, prove that the crust was rigid and cool prior to mafic dyke intrusion; this dilatational event is named as Eₐ. The mafic dykes were metamorphosed and altered in a potassium-metasomatic event, Eₐ, that was progressively developed to high-temperature, penetrative shortening accompanied by production of garnet-bearing granitic veins and dykes. The Eₐ-Eₐ events are related to an progressive extension that was followed by contractional steep folding events. Isoclinal Fₐ characterizes the outcrop and surrounding area. Large-scale asymmetric transpressional folding, Fₐ with a NE-SW-trending axial plane deformed the earlier structures. The Eₐ event was followed by angular Fₐ folding with N-trending axial planes suggests c. E-W contraction. An ESE-WNW-trending coarse-grained pegmatitic granite dyke, which yielded a U-Pb monazite age of 1804 ± 2 Ma, cuts the D₉ structures as shown by gneiss fragments including Fₐ folds in the dyke. A pegmatitic granite dyke was intruded into a dilatational fracture after or during the latest stage of the Eₐ event, partly regulated by ductile sinistral N-S shearing near the dyke contacts. The latest structures, D₉-D₉, are open warps of the early structures. Fₐ folds are recumbent warps that were refolded by the open Fₐ folds formed under N-S compression. These deformations did not markedly affect the pre-existing structures on outcrop scale, but they are important when interpreting the regional variations in metamorphic grade, geophysical anomaly patterns or continuation of lithological sequences on the present-day erosion surface.

Key words (GeoRef Thesaurus, AGI): structural geology, gneisses, granites, dikes, structural analysis, deformation, absolute age, U/Pb, Proterozoic, Paleoproterozoic, Västerviken, Karjalohja, Finland

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INTRODUCTION

The c. 1.9 Ga old Svecofennian domain in the central part of the Fennoscandian Shield collided with the Archaean continent in the east at 1.89–1.885 Ga ago (Wegmann 1928, Koistinen 1981 and Korsman et al. 1999) (Figure 1). The multi-phase accretion, collision and extension history (Koistinen 1981, Lahtinen 1994, Korsman et al. 1999, Kilpeläinen 1998, Pajunen et al. 2004 and 2008) led to the recent reconstructions of geotectonic terranes each with individual characteristics representing different primary formation environments, timing and tectonomagmatic histories. The 1.93–1.92 Ga primitive Pyhäsalmi island arc (Korsman et al. 1984, Lahtinen 1994, Korsman et al. 1999) was fragmented during collision against the craton boundary (Koistinen 1981 and Koistinen et al. 1996). The central Finland granitoid complex (cf. Nironen 2003) covers a wide area and is composed of tectonic slices stacked towards the N-NW (Pajunen et al. 2008) between the primary arc and the more mature arcs in the west and south. It shows a polyphase magmatic history with large amounts of syntectonic, mostly I-type granitoids (Nurmi & Haapala 1986) that were then intruded by post-kinematic granites that are younger towards the west (Rämö et al. 2001). In the south younger, c. 1.9–1.88 Ga old (Vaasjoki & Sakko 1988 and Väisänen & Mänttäri 2002), volcanic-sedimentary belts trend roughly E-W and show a primary accumulation environment with more deepwater characteristics towards the north (Kähkönen 1987 and Ehlers & Lindroos 1990). Southern sequences were intruded by huge amounts of felsic granitoids that form the c. 100 km wide ENE-trending Southern Finland Granite Zone, SFGZ (Figure 1). Major high-strain or polyphase shear and fault zones separate the terrains (Pajunen et al. 2008).

As early as 1926 J. J. Sederholm described a tectono-magmatic discordance with cooling stages between the magmatic and migmatitic pulses in migmatitic granitoids in the southernmost Svecofennian terrain in the Tammissaari archipelago. Korsman et al. (1984 and 1988) established a corresponding tectono-metamorphic discordance between the Trondhjemite Migmatite Belt (TMB) and Granite Migmatite Belt (GMB) (Korsman et al. 1997 and Figure 1). The crust largely stabilized in the north c. 1.875 Ga ago, whereas in the south thermal heat flow remained high still later at c. 1.86–1.80 Ga ago (e.g. Korsman et al. 1999 and Pajunen et al. 2008). Several detailed or regional structural analyses to determine the complex evolution of the Svecofennian domain have been published during the last 30 years (e.g. Laitala 1973, Hopgood et al. 1976, Verhoef & Dietvorst 1980, Koistinen 1981, Hopgood 1984, Ehlers et al. 1993, Koistinen et al. 1996, Korsman et al. 1984, 1988 and 1999, Kilpeläinen 1988 and 1998, Kilpeläinen & Rastas 1990, Pietikäinen 1994, Nironen 1999, Pajunen et al. 2001, Väisänen et al. 1994, Väisänen 2002, Väisänen & Hölttä 1999, Pajunen et al. 2004 and 2008). These analyses give a blanket overview of the evolution in the domain, but there are many contradictions in details, especially, in the southernmost parts. Correlation of deformation events over the major tectono-metamorphic discordances (e.g. Korsman et al. 1984) has been difficult. Correlation problems arise, because different tectono-thermal terranes display dissimilar "steps" in the deformation succession (Hopgood et al. 1976 and Pajunen et al. 2008). On the present-day erosion level slices from different crustal depths are exposed, for instance the Orijärvi area (Eskola 1914 and Kilpeläinen & Rastas 1990) contrasts with the neighbouring granulite facies West Uusimaa Complex (Parras 1958, Shreurs & Westra 1986 and Kilpeläinen & Rastas 1990). Accordingly, structures developed simultaneously in various depths exist in a close association to each other. In many cases they differ in appearance and tectonic style due to different ductility contrasts at various depths. Kilpeläinen (1998) evidenced that some early structures are totally missing from the tectonic slices representing the upper crustal levels during deformation.

The main goal of this paper is to provide data from a well-studied structural succession integrated with two new age determinations from a key outcrop as a contribution to current attempts to better establish the tectonic evolution of the southern Svecofennian crust. The key outcrop, in the Västerviken, Karjalohta (Figures 2 and 3), is situated in a migmatitic terrane, to the southeast of the Orijärvi area that was described in the classical papers of Eskola (1914 and 1915). Felsic gneisses, the origin of which has been under debate since Eskola’s time, characterize the terrane. Between felsic gneisses there are zones of intermediate to mafic volcanic rocks and pelitic to psammitic metasedimentary gneisses and calc-silicate rocks. Several pulses of intrusive rocks and migmatization veins intrude the supracrustal successions (Sederholm 1926 and Hopgood et al. 1983).

For the present study T. Koistinen and A. Hopgood carried out the primary fieldwork on the outcrop in the early 1990’s. Samples for isotopic dating, carried out by H. Huhma, were collected at 1991. The authors collaborated in the spring of 2004 to review the data, in thin section and on the outcrop, and integrated this with the current knowledge on the tectonic evolution of the Svecofennian domain.
Integrated structural succession and age constraints on a Svecofennian key outcrop in Västerviken, southern Finland

Figure 1. Study area (star) in Västerviken Karjalohja (Finnish grid coordinates KKJ2 x = 6674300 and y = 24860900) on the geological map of southern Finland. Simplified from Korsman et al. (1997). SFGZ = Southern Finland Granite Zone, O = Orijärvi area, T = Tammisaari area, TMB = Tonalite Migmatite Belt (trondhjemite migmatite belt) and GMB = Granite Migmatite Belt.
Figure 2. Tectonic setting of the studied Västerviken outcrop.

Figure 3. Detailed tectonic structure of the Västerviken outcrop.
DESCRIPTION OF THE VÄSTERVIKEN OUTCROP

Lithology

The main rock type on the Västerviken outcrop is felsic gneiss (Figures 4 and 5) that was intruded by intermediate to mafic dykes and several pulses of felsic veins and dykes. The felsic gneiss shows variably distinguishable banding that at least partly represents the original bedding, S₀. Often S₀ is destroyed by polyphase deformation and concomitant metamorphic recrystallization, but when visible, it exhibits as minute variation of grain-size or composition. Compositional layering appears as biotite-, biotite-green hornblende-rich, amphibolitic and calc-silicate-bearing layers. Intermediate to mafic layers in the felsic gneiss imply a volcanic or volcanoclastic-sedimentary origin for the rocks of the Västerviken outcrop. Under the microscope the layered felsic gneiss is fine-grained, the grain size being on average less than 1 mm, and granoblastic composed of plagioclase, quartz and biotite, with opaque minerals, apatite and zircon as minor minerals. Small amounts of up to 1 cm subidioblastic or xenoblastic garnet grains exist in the biotite-rich layers (Figure 6a) or in flattened biotite patches in the more homogeneous parts. Some garnets show a two-stage growth with quartz and some biotite inclusions with vague internal foliation in their cores (I in Figures 6b and c) that are overgrown by rims including relic biotite inclusions (II in Figure 6c). The primary origin of the homogeneous felsic gneisses has been under debate for years and at present there exists no widely accepted consensus. The host gneiss is bordered by more homogeneous tonalitic gneiss along tectonic zones (1 in Table 1). In contrast to the banded gneiss, some potassium feldspar occurs in homogeneous parts of the rock. A sample of the layered felsic gneiss was chosen for isotopic study (sample A1295, Table 2).

The felsic gneiss is cut by intermediate to mafic dykes (5a in Table 1) up to 0.5 m wide (Figure 7) that are strongly deformed and altered by later events. Isoclinally folded felsic segregation veins (6 in Table 1) indicate intense shortening in the dykes (Figure 8). Rarely, in the least deformed parts near fold hinges, a vague ophitic structure can be identified; one dyke shows plagioclase-porphyrritic structure (5b in Table 1). The dyke cuts sharply across the felsic gneiss and the early phase of felsic neosome and pegmatitic veining (Figure 7); thus it post-dates the early migmatization pulses and boudinage of the felsic segregation veins. In the better-preserved parts the dyke is composed of hornblende, plagioclase and quartz with minor opaque minerals. Mostly the dykes are penetratively altered and granitized resulting in breakdown of almost all the hornblende to biotite; alteration occurs in zones parallel to the dyke. In these tightly folded dykes quartz often shows undulose extinction and biotite shows intense crenulation cleavage.

Several generations of neosome-forming felsic vein and dyke material can be distinguished on the outcrop. The early narrow quartz-feldspar segregation veins (2 in Table 1) and the isoclinally or pytgmatically folded, fine- to medium-grained trodhamitic
Figure 4. $F_a$ folding in felsic banded gneiss. Photo by T. Koistinen.

Figure 5. $S_a$ foliation deformed by $F_b$ folding in homogeneous weakly banded felsic gneiss. Photo by T. Koistinen.
Figure 6. a. Two-stage garnet grains in banded and folded felsic gneiss. Rims of the grains overgrow $S_b$ foliation. Photo T. Koistinen.
b. Two-stage garnet with a vague internal $S_b$ foliation in the core (I) and nearly inclusion-free rim (II) overgrowing the $S_a$. c. Garnet rim (II) overgrowing the $S_b$ biotite. Photos by M. Pajunen.
Figure 7. Mafic dyke cutting a pinch and swell structured, D_1 boudinaged, pegmatitic granite dyke (4 in Table 1). The cutting relationships indicate cooling after D_1. Photo by T. Koistinen.

Figure 8. Isoclinally folded felsic segregation vein (6 in Table 1) in a mafic dyke (5a in Table 1) indicating strong shortening, D_2, of the dyke. Photo by M. Pajunen.
veins (3 in Table 1) migmatize the host gneiss (Figure 9). Narrow biotitic selvages with minute garnet grains border the veins; garnet has not been found in the veins themselves.

A later generation of more coarse-grained vein- ing (4 in Table 1) is also isoclinally folded and changes from trondhjemitic to granitic in composition (Figure 7 and 10). No garnet was found in the veins. They show biotite-rich rims against the host rock. These dykes form ductile pinch and swell structures that locally form totally isolated swells; in places, only the border zone of the dykes shows pinch and swell structure. The spaced foliation of surrounding gneiss parallels the contacts of the pinch and swell structured pegmatite veins. The mafic dykes described above cut these felsic veins (2–4 in Table 1).

Foliated pegmatitic granite dykes (7a in Table 1) overprint the early felsic veining (2–4 in Table 1) (Figure 10). In contrast to the previous ones they include sparse, up to 1 cm garnet grains. The pegmatitic granite dykes show tight folding and bulging of the borders with some quartz in the shallow boudinage necks. However, strain including boudinage was not as intense as in the pinch and swell dykes (4 in Table 1). There are weak biotite melanosome on the dyke borders, but between the isoclinally folded limbs of folded dykes, biotite was coarsened and indicates metasomatic potassium segregation during the flattening process. The borders of the dykes vary between being sharply penetrating to more or less gradational against the earlier vein generations. In detail these garnet-bearing dykes cut the pre-existing veins (2–4 in Table 1) and the foliation related to them (Figure 11). Vague coarse-grained granitic patching and narrow, ptygmatically folded veining (7b in Table 1) of granitic composition similarly overprint the earlier felsic migmatization veins.

The latest coarse-grained pegmatitic granite dykes (8 in Table 1) (Figure 12) cut all the described structures except for those classed as E₀ and E₁ (Table 1). The dykes are up to five metres wide and c. ESE-WNW-trending, dipping 65° to the south. In the northern contact zone of the widest dyke the granite is medium-grained and changes to coarse-grained towards the central part; in the southern contact the rock is coarse-grained. The contacts of the dykes are quite sharp, but showing slight pinch and swell, indicating some ductile to semi-ductile deformation during their intrusion. Their tectonic
setting can be placed from enclaves up to 0.5 m of the deformed country rock that contain pre-existing $F_8$ folds (Table 1). Locally, the contacts are sinistrally sheared in a N-S direction. However, the coarse biotite shows no determinable foliation in the central parts (Figure 13). The dykes are potassium feldspar-rich, quartz occurs up to 20 cm wide patches and patchy biotites have grown up to 15 cm grains; no garnet was found. The dykes are widespread in the surrounding outcrops indicating an important magmatic pulse. This rock was chosen for isotopic dating (sample A1297, Table 2).

**Deformation succession and its regional correlation**

The following description of the deformation succession (Table 1) exposed at Västerviken concentrates on the local major structures that can be identified on the outcrop (Figures 2 and 3). Not all of the minor features of the detailed field study are described here, because their regional significance has not yet been ascertained, and some structures observed in regional scale are not represented in the studied outcrop (cf. Pajunen et al. 2008).

**$E_a$ event**

The earliest deformation structures identified on the outcrop are small isoclinal, intrafolial folds $F_a$ that deform the primary layering $S_0$. The fold limbs have become intensely thinned and their hinges are sharp and attenuated (Figure 4). Relict $S_a$ is identified as a weak biotite foliation bending around $F_a$ fold hinges in the felsic gneiss and weak felsic segregation (Figure 5). Elsewhere its identification is uncertain. The early folds are not identified in the non-layered tonalitic gneisses (1 in Table 1). There are also some quartz-(feldspar) veins (2 in Table 1) post-dating $S_a$ (Figure 9). Metamorphic assemblages during $D_a$ cannot be established with certainty, because later higher-grade events have overprinted them; $S_a$ biotite/hornblende foliation was accentuated by $D_a$ mineral growth where $S_a$ is parallel to it (Figure 14) and there is a mineral lineation, $L_a$, parallel to $F_a$ fold axes. The establishment of early garnet growth...
Figure 11. Detail from the contact of a garnet-bearing granite dyke (7 in Table 1). The dyke cuts both the veins of earlier migmatization pulses and the $S_b$ foliation. Photo by M. Pajunen.

Figure 12. Pegmatitic granite dyke (8 in Table 1) cutting across a large F$_f$ fold, which shows as a bend in the $S_b$ foliation and the F$_f$ fold axes. Fr = fragment of country rock and red box = sample A1297. Photo by M. Pajunen.
with respect to deformation is tentative, but some curved inclusion trails in garnet cores implies early growth (Figure 6b). If the vague internal foliation in garnet cores is a relic \( S_a \), then the biotite-quartz-garnet assemblage is a result of the early prograde metamorphism, \( M_p \), to amphibolite facies during \( D_b \) similar to that described by Hopgood et al. (1976) from c. 30 km southwest of the study area, in Skåldö, Tammisaari. The \( F_b \) fold axis is nearly horizontal, plunging gently east on the outcrop; estimation of tectonic transport direction is only tentative due to reorientation of \( F_a \) by later deformation.

**Eb event**

\( D_b \) deformation was associated with the tectonic episode that produced the regional foliation, segregation and migmatization that characterizes the outcrop. \( D_b \) was accompanied by intense shortening as shown by small-amplitude, isoclinal to tight, intrafolial folds, \( F_b \), that refold the \( S_b \) foliation (Figure 5). \( S_b \) is axial planar mineral foliation, penetrative biotite foliation or spaced crenulation and segregation of coarse- and fine-grained bands in felsic gneisses. \( F_b \) hinges are typically round and locally shearing occurred along the fold limbs promoted by the increased ductility resulting from continuing melting. In the layered gneiss the \( F_b \) folds are generally tight and less than 1 m in amplitude and \( F_b \) fold axes plunge gently approximately E or W. \( S_a \) and \( S_b \) form an intersection lineation (\( L_{a/b} \)) that is best seen in the felsic gneiss when \( S_b \) forms a strongly recrystallized spaced biotite crenulation intersecting the \( S_b \) segregation foliation. The early plagioclase-quartz segregation veining (2 in Table 1) was produced along \( S_b \) planes and anatctic pulses progressed to more intense trondhjemitic neosome (3 in Table 1) formation. These early veins are isoclinally folded and attenuated indicating intense flattening strain (Figure 9). The later, more coarse-grained granite veins (4 in Table 1) have spaced \( S_b \) foliation lying parallel to the vein contacts (Figure 10) and were deformed into boudins and pinch and swell structures (Figure 7). Garnet rims (II in Figures 6a and 6b) overgrow \( S_b \) that is preserved as some \( M_b \) biotite relics (Figure 6c), but \( S_b \) is curved to surround the garnet core (I in Figure 6b).
Table 1. Succession of structural elements on the Västerviken outcrop.

<table>
<thead>
<tr>
<th>Event</th>
<th>Structure</th>
<th>Description</th>
<th>Process</th>
</tr>
</thead>
<tbody>
<tr>
<td>E₀</td>
<td>S₀</td>
<td>Primary layering, sedimentary or volcanic (Figure 4).</td>
<td>Folding of indurated sedimentary layers.</td>
</tr>
<tr>
<td>E₁</td>
<td>F₁</td>
<td>Intrafolial, isoclinal, sharp-hinged folds with E-trending axis (Figure 4).</td>
<td>Onset of T increase.</td>
</tr>
<tr>
<td></td>
<td>S₁</td>
<td>Axial plane penetrative mineral foliation (strengthened by M₁) or vague internal foliation in garnet (Figure 6b).</td>
<td>Low-grade metamorphism up to amphibolite facies.</td>
</tr>
<tr>
<td>L₁</td>
<td></td>
<td>Mineral lineation (predominantly Hbl at present, Figure 14), intersection between S₁ and S₂.</td>
<td>Early-Ds segregation veining – low-T (cf. Mouri et al. 1999).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Tonalites (1), approximately concordant to S₁, includes S₂.</td>
<td>Intrusion into the supracrustal gneisses.</td>
</tr>
<tr>
<td>E₂</td>
<td>F₂</td>
<td>Intrafolial, tight to isoclinal, round hinge, refolds S₁ (Figure 5).</td>
<td>Cooling. Crust becomes rigid.</td>
</tr>
<tr>
<td></td>
<td>S₂</td>
<td>Axial plane penetrative mineral foliation (Hbl, Bt) (e.g. Figures 5, 6 and 7).</td>
<td></td>
</tr>
<tr>
<td>I₁</td>
<td></td>
<td>Qtz-Fsp segregation veins (2, Figure 9), parallell to S₂, cut by veins (3); relation to tonalite (1) not seen.</td>
<td></td>
</tr>
<tr>
<td>I₂</td>
<td></td>
<td>Trondhjemitic veins (3), strongly folded and boudinaged (Figure 9).</td>
<td></td>
</tr>
<tr>
<td>I₃</td>
<td></td>
<td>Granitic veins (4), folded, pinch and swell structured and boudinaged, compositional difference between veins (3) and (4), folds rounder in vein (4) than in vein (3) (Figures 7 and 10).</td>
<td></td>
</tr>
<tr>
<td>E₃</td>
<td>F₃</td>
<td>Intrafolial, small, tight angular folds with sharp hinges in altered dyke (Figure 15) isoclinal round hinges in folded segregation veins (6) (Figure 8).</td>
<td>T increase. Flattening perpendicular to foliation.</td>
</tr>
<tr>
<td></td>
<td>S₃</td>
<td>Spaced foliation.</td>
<td>Low-T biotitic alteration in some mafic dykes (5).</td>
</tr>
<tr>
<td>L₃</td>
<td></td>
<td>Intersection lineation L₃ near the dyke (5a) contact folded by F₃ (Figure 18).</td>
<td>T increase and partial melting.</td>
</tr>
<tr>
<td>I₃</td>
<td></td>
<td>Granite veins (7a), garnet-bearing, cuts veins (3) and (4) (Figures 10 and 11), less deformed (flattened) than dykes (3) and (4).</td>
<td></td>
</tr>
<tr>
<td>I₄</td>
<td></td>
<td>Pyrgmatig granite veins and melt patches (7b).</td>
<td></td>
</tr>
<tr>
<td>E₄</td>
<td>F₄</td>
<td>Large amplitude (more than 5 m) isoclinal folds (Figure 16). Refolds S₃.</td>
<td>Intense compression.</td>
</tr>
<tr>
<td></td>
<td>S₄</td>
<td>Thinned limbs. Relation to dykes (7a) not seen.</td>
<td>Some biotite growth into S₃ axial plane (Figure 17).</td>
</tr>
<tr>
<td>L₄</td>
<td></td>
<td>Intersection lineation between S₃ and S₄ parallel to hinges (Figure 18).</td>
<td></td>
</tr>
<tr>
<td>E₅</td>
<td>F₅</td>
<td>Large (amplitude tens of metres) moderate folds (Figures 2, 3 and 12). Axial planar trend of F₅ is c. NE (Figures 2 and 3). Steep fold axis.</td>
<td>Moderate horizontal SE-NW compression.</td>
</tr>
<tr>
<td></td>
<td>S₅</td>
<td>Weak local spaced axial planar cleavage approximately 160/90 (Figure 19).</td>
<td></td>
</tr>
<tr>
<td>E₆</td>
<td>F₆</td>
<td>Sub-angular moderate folds (Figure 20).</td>
<td>Moderate horizontal E-W compression.</td>
</tr>
<tr>
<td></td>
<td>S₆</td>
<td>Axial plane approximately 090/90.</td>
<td></td>
</tr>
<tr>
<td>I₆</td>
<td></td>
<td>Coarse-grained pegmatitic granite dykes (8) dated at 1804 ± 2 Ma. Contacts c. 180/65. Local N-S-shearing in the contacts. Includes enclave of F₅ fold.</td>
<td></td>
</tr>
<tr>
<td>E₇</td>
<td>F₇</td>
<td>Open recumbent folds (Figure 16) refold S₆ axial plane.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>S₇</td>
<td>No axial plane structure observed. Axial plane dips gently north.</td>
<td></td>
</tr>
<tr>
<td>L₇</td>
<td></td>
<td>Wrinkling of pre-existing biotitized zones along the fold hinge (approximately 270/10).</td>
<td></td>
</tr>
<tr>
<td>E₈</td>
<td>F₈</td>
<td>Open warps refold S₇ (Figure 16).</td>
<td>Weak horizontal N-S compression.</td>
</tr>
<tr>
<td></td>
<td>S₈</td>
<td>Axial plane approximately 000/90.</td>
<td></td>
</tr>
</tbody>
</table>
The Ec mafic dykes (5a in Table 1) cut the foliation, trondhjemitic veining (3 in Table 1) and also split the pinch and swell structure of the early pegmatite dykes (4 in Table 1). Some faulting occurred during the dyke emplacement because of non-continuation of the pinch and swell structure on the opposite side of dyke (Figure 7). Their sharp and originally planar contacts suggest intrusion into cooled, rigid crust indicating a stabilizing and cooling stage after the Eb event. The plagioclase-porphyritic dyke (5b in Table 1) shows a similarly retrogressed (see below) mineral assemblage suggesting its emplacement corresponded to the mafic dyke. The deformation structures in the dykes show post-Ec structural history.

Ed event

Retrograde, spaced foliation (Figure 15) in the mafic dykes is Fd-folded (Figure 15 and 16); under the microscope some remnants of small tight angular Fd folds are crenulated by De. Locally the hornblende-plagioclase assemblage of the dyke was altered to almost monomineralic coarser-grained (3 mm) biotite (Figure 17). This type of alteration implies metamorphic potassium enrichment during the development of the foliation. In the strongly altered dyke contacts composite foliation $S_{a+b}$ and the spaced foliation $S_d$ formed an intersection lineation $L_d$ ($L_{(a+b)/d}$ in Figure 18).

Narrow, isoclinal-folded feldspathic segregation veins (6 in Table 1) cut the mafic dyke (Figure 8); originally the veins formed approximately perpendicular to the dyke. The shortening in the garnet-bearing granite dykes (7a in Table 1) (Figures 10 and 11) corresponds to that in the mafic dykes; they intruded prior to Ec. Isoclinal folding in these dykes (7a in Table 1) (Figure 2) and tight folds in felsic gneiss (Figure 9) indicate that the continuing Dd deformation under increasing temperature affected the rock more thoroughly. Deformation occurred under changing conditions of potassium-rich fluidal activity during the granitic anatectic event. The formation of these garnet-bearing granite dykes, more irregular granitic melt patches and ptygmatic veins (7b in Table 1) indicates a new thermal pulse with temperature increase to high-grade conditions related to flattening strain.

Ee event

Ductile Fz folds in the mafic dykes are tight with amplitudes of several metres (Figures 2, 3 and 16). Their hinges are relatively thicker than their limbs.
Figure 15. Photomicrograph of an intermediate dyke showing zoned alteration of hornblende (Hbl) to biotite (Bi). The altered zones are isoclinally $F_d$ folded and they are further crenulated in $D_c$. Photo by M. Pajunen.

Figure 16. Isoclinally $F_e$ folded mafic dykes (5a in Table 1). The axial plane $S_e$ was bend by later open recumbent folding $F_h$ (axial planar attitude $000^\circ/45^\circ$) and c. N-S trending compression $D_c$. Photo by T. Koistinen.
Figure 17. Microphoto of S_c crenulation in penetratively biotitized (D_d) mafic dyke. Photo by T. Koistinen.

Figure 18. L_e intersection lineation folded around an F_e fold. The intersection lineation L_e between the biotitic alteration-segregation foliation S_d and S_e is parallel to the F_e fold axis. Photo by M. Pajunen.
In the felsic gneiss $F_e$ are rounder and fragmented reflecting competence contrast between the dyke and gneiss. $S_e$ foliation dips steeply south. $F_e$ fold axes are bent by later deformation and have a shallow plunge. $S_e$ biotite foliation was generated in the most intensely folded hinges. Generally, in the mafic dykes, $S_e$ is a crenulation cleavage with some biotite in the axial plane (Figures 15 and 17). The intersection lineation, $L_s$, was bent around the $F_e$ hinge (Figure 18). The angular disparity between $F_e$ hinges and the lineation serves to separate them as products of different deformation episodes, indicating different tectonic transport directions with respect to the early deformation. In the dykes themselves the $S_e$ alteration segregation foliation bends around the $F_e$ fold hinges. The $S_d$ and $S_e$ form a weaker intersection lineation $L_e$ that is parallel to the $F_e$ fold axis. The two intersection lineations can be clearly distinguished on the outcrop (Figure 18).

$E_f$ event

$D_f$ deformed the $S_e$ axial surfaces into semi-open, dextral, asymmetric folds with c. NE-trending axial planes, $S_f$ (Figures 2 and 3). On the outcrop $F_f$ axes can be seen to plunge steeply (Figure 12), but in areas characterized by horizontal earlier structures, the plunge is typically shallow (Pajunen et al. 2008). The amplitude of $F_f$ folds ranges between tens of metres and kilometres, and the folds are clearly visible at map scale, for example on aeromagnetic and lithological maps (1:100 000). On the Västerviken outcrop weak, spaced $S_f$ cleavage deforming the main foliation in the limbs of $F_f$ folds can sometimes be observed (Figure 19).

$E_g$ event

Sub-angular moderate folds $F_g$ (Figure 20) with N-trending steep axial planes refold the $F_f$ fold limbs and suggests E-W contraction. No $S_g$ axial plane foliation was observed, but local N-trending sinistral shearing along the granite pegmatite dyke (8 in Table 1) occurred.

The granite pegmatite dyke (8 in Table 1) cuts all the structures described above (Figures 2, 3 and 12). It includes fragments of $F_g$-folded country rock. The dyke intruded into a dilatational fracture in c. ESE-WNW direction and is related to the latest phase of $E_g$ or shortly post-dating it.
 Eh, Ei and other late events

$S_h$ axial planes are curved to form open recumbent folds with shallow $S_h$ axial planes dipping north (Figure 16). No axial planar foliation was observed, but there are $F_h$ wrinkles in the pre-existing biotitized zones with approximately 270°/10° axes.

Very open $F_i$ warps refold the $S_h$ axial surface (Figure 16). The $S_i$ axial plane is steep and approximately E-trending indicating N-S contraction. In the studied outcrop it cannot be established whether the lower-grade, semi-ductile deformation zones with fine-grained, neomineralized biotite and deformed quartz, overprinting the higher-grade $E_i$ kinking in biotite, is related to these late open warps. These microstructures indicate that deformation took place under semi-brittle conditions, characteristics that are related to the latest Svecofennian, and at least partly, to post-Svecofennian events by Elminen et al. (2008) and Mertanen et al. (2001 and 2008).

Isotopic results

Mineral separation of the felsic gneiss A1295 yielded a small amount of small, short, nearly colourless and transparent crystals of zircon. Many grains have slightly rounded edges due to resorption. The zircon population appears to be homogeneous, and consistent with a felsic metavolcanic rock. The U-Pb data are shown in Table 2 and Figure 21. The two U-Pb analyses on zircon from the felsic gneiss sample A1295 are discordant and do not allow a reliable age determination. The analyses plot on the chord with intercepts at 1881 and 291 Ma.

The pegmatitic granite (8 in Table 1) A1297 contains monazite, which in the extracted fraction occurs mostly as turbid and reddish fragments. The heavy minerals also contain a small amount of low-density zircon, which was not analysed in this study. In spite of the high content of U, the U-Pb analysis on monazite from the pegmatite A1297 is nearly concordant and provides an age of 1804 ± 2 Ma (Figure 21). This is considered to be an intrusive age for the pegmatite generation, which postdates the major ductile folding deformations ($F_{a-g}$) observed in the outcrop.
Integrated structural succession and age constraints on a Svecofennian key outcrop in Västerviken, southern Finland

Table 2. Mineral U-Pb data from the Västerviken samples.

<table>
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<tr>
<th>Sample information</th>
<th>Sample</th>
<th>U (mg)</th>
<th>Pb (ppm)</th>
<th>206Pb/238U</th>
<th>207Pb/235U</th>
<th>208Pb/206Pb</th>
<th>ISOTOPIC RATIOS*</th>
<th>Rho**</th>
<th>APPARENT AGES / Ma</th>
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<td>A1295 felsic metavolcanic rock:</td>
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<tr>
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<tr>
<td>A1297A Monazite</td>
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<td>9477</td>
<td>11641</td>
<td>2.04</td>
<td>0.318</td>
<td>1</td>
<td>4.84</td>
<td>1</td>
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</tbody>
</table>

*)Corrected for blank (Pb=0.5 ng, U=0.2 ng), mass fractionation and initial age related common lead (Stacey & Kramers, 1975).
**)Error correlation
Zircon in A1295: small, short, light, translucent

Figure 21. Concordia diagram of zircon and monazite analyses from the Västerviken samples.
DISCUSSION

Deformation

The Västerviken outcrop represents a polyphase and migmatitic equivalent of the Svecofennian amphibolite facies supracrustal gneiss terrane southeast of the Orijärvi area. The structural sequence on the outcrop yields two major ductile events, $E_2$-$E_3$ and $E_4$-$E_5$. Semi-brittle to brittle fracturing channeling the mafic dyke magmatism, $D_c$, indicates a cooling stage between the ductile events. Late ductile to semi-ductile open folding phases $E_{h-i}$ do not show any marked effects on the outcrop scale, but they are important when in the context of map-scale structures and variations in regional metamorphic assemblages (cf. Pajunen et al. 2008).

$D_b$ was a strong vertical shortening event with concurrent felsic neosome formation; trondhjemite veins (3 in Table 1) and pinch and swell structured veins (4 in Table 1). The $D_b$ structures were deformed into steep position during later deformation events (e.g. Pajunen et al. 2008). Pajunen et al. (2004) related the corresponding structures to a crustal collapse, and later, to an island arc collapse (Pajunen et al. 2008). Unfortunately, no acceptable material was obtainable for conventional U-Pb analyses from the granitic pinch and swell veins in Västerviken (sample A1296). Many age determinations provide evidence for an early migmatitic-metamorphic history in the vicinity of Västerviken. According to Hopgood et al. (1983) large euhedral sphenes from a quartzofeldspathic vein (A798 Kurö) crosscutting banded dioritic gneiss provide concordant analysis with an age of 1882 ± 9 Ma. This vein shows weak effects of $F_{c}$ folding of Hopgood et al. (op. cit.) and the date is supported by concordant 1877 ± 6 Ma old monazite from garnetiferous gneiss (A806). Pink aplite veins containing monazite with a concordant age of 1871 ± 3 Ma (A797-Orrholman) cut the quartzofeldspathic veins. An example from trondhjemite migmatites (cf. Korsman et al. 1984) is provided by the study of Mouri et al. (1999), who report ages of c. 1.88 Ga for monazite and...
garnet in Luopioinen, c. 50 km ESE from Tampere. In the present outcrop, the main garnet growth (inclusion-free rims) post-dated \( D_p \), but the \( S_b \) foliation bends around the garnet core with inclusion trails. Correlation between the ages and structures described indicate that \( D_p \) occurred c. 1.88 Ga ago, at about the same time as the youngest volcanic rocks in the Orijärvi area were emplaced. Thus, several pulses of felsic vein formation occurred over a very short time indicating a c. 1.88 Ga magmatic event with concomitant volcanism (cf. Väisänen & Mänttäri 2002) in the Southern Finland Granite Zone (Figure 1).

In Västerviken the cooling stage after \( E_s \) is evidenced by the intrusion of the mafic dykes, \( M_e \), into a semi-brittle to brittle fractures of the \( E_s \)-migmatized host rock, and by the formation of the retrograde, metasomatic segregation foliation \( S_d \) into the mafic dykes. Earlier in 1926 Sederholm described a similar discordance from the Tammisaari archipelago. Several studies (e.g. Korsman et al. 1984) divided the crustal evolution in the southernmost Svecofennian into two major tectono-thermal stages, the c. 1.9–1.88 Ga and c. 1.83–1.80 Ga events. The earlier stage was followed by felsic magmatism and fluidization over wide areas of the Svecofennian domain, and even in the eastern part of the Archaean terrain at c. 1.85 Ga ago (compiled in Pajunen & Poutiainen 1999). Recently, several age determinations of magmatic rocks (e.g. Kurhila et al. 2005) and of tectono-metamorphic processes have widened the age range of magmatic and metamorphic processes to cover the whole 1.86–1.80 Ga evolution (e.g. Pajunen et al. 2008). The \( E_e \) mafic dykes marking the tectonic break in Västerviken were not dated but the granitic neosome veins (7a-b in Table 1) postdating the \( E_e \) dykes provide evidence of new metamorphic progression. Intermediate dykes in a corresponding setting to the mafic dykes are common in the southern Svecofennian area (e.g. \( D_b \) microtonalite dykes of Pajunen et al. 2008). They indicate intrusion into cooled crust and were later migmatized by a granitic migmatization pulse, except for in the granulite areas, where heat flow remained continuously high during these late stages of the Svecofennian evolution.

The granitic veins (7a in Table 1) are \( F_q \)-folded, but are not as strongly deformed as the typical \( D_k \) veins (3 an 4 in Table 1). The structural setting of the granitic dykes (7a in Table 1), patchy pegmatitic granite neosome and narrow ptygmatic granite veins (7b in Table 1) corresponds to the granite-pegmatite dykes and contemporaneous veining (\( D_q \)) described by Pajunen et al. (2008) from Bollstad, Inkoo, c. 18 km SE from Västerviken, but are deformed into steep position. In Bollstad the pegmatitic granite dyke intruded into a brittle fracture that is related to the onset of oblique extension (\( D_j \)). Thus, the age of the pegmatitic dyke fixes the beginning of the episode in the area at 1828 ± 3 Ma ago (Pajunen et al. 2008). As the extension evolved, granitic migmatization increased and finally large amounts of late-Svecofennian granites were intruded. Continuing vertical shortening with increasing heat flow signified the extensional deformation (\( D_i \) of Pajunen et al. 2008).

The latest data on tectonic and magmatic events establish the Svecofennian evolution as a more continuous process under regional transpressional deformations that began at c. 1.87–1.86 Ga ago and continued under varying stress field to c. 1.8 Ga (Pajunen et al. 2008). Earlier Ehlers et al. (1993) represented the first postulations on transpressional evolution in the Svecofennian domain; they interpreted the intrusion of the 1.84–1.83 Ga granites as synkinematic to the transpressional event at 1.83 Ga. These transpressional deformations caused alternating oblique extensional and contractional structures that caused varying structural sequences in different tectonic units or crustal depths. According to Pajunen et al. (2008), the SE-NW extensional process (\( D_p \)) was closely followed by N-S contraction causing wide-scale folding with E-W-trending axial planes (\( D_q \)). It was followed by transpression in a NE-SW direction (\( D_h \)), and during the latest Svecofennian stages the stress field turned to c. E-W-directed (\( D_j \)). The high thermal activity related to these transpressional processes show the most continuous character in the granulate areas of southernmost Finland. The U-Pb results on monazites from the West-Uusima granulites c. 10 km north from Västerviken also provide ages of 1.83–1.82 Ga, whereas Sm-Nd analyses on garnet suggest that the granulites cooled to garnet Nd-blocking temperature (c. 650–700 °C) at about 1.8 Ga ago (Väisänen et al. 2004). Väisänen et al. (2002) have reported U-Pb sims results on zircons from the Turku area. These include an age of 1826 ± 5 Ma for the leucosome of the Lemu garnet-cordierite gneiss, and ages of c. 1.8 Ga for the leucosome of the Masku tonalite and for the metamorphic zircon rims in the 1878 ± 19-Ma-old Kakskerta enderbite.

In the Västerviken outcrop the coarse-grained 1804 ± 2 Ma old pegmatitic granite (8 in Table 1) (A1297) post-dates the major ductile deformation fabrics. It contains fragments of the host rock with \( F_q \) folds that exhibit c. E-W contraction. The dyke contact shows a sinistral N-S shear component implying local (?) rotation of the stress field after the E-W-contractional \( E_e \) event into a more NW-SE trend. Rotations of stress field during the late deformations, \( D_q \) (\( D_i \) of Pajunen et al. 2008), postulate
variations in setting of the extensional granitic magmas during the heterogeneous heat flow and deformation from c. 1.84 to 1.80 Ga. The age of 1.8 Ga is common throughout the Shield. The two medium-grained pink aplogranites in Tammisaari (see Hopgood et al. 1983), which post-date folds of FE generation of Hopgood et al. (op. cit.) gave discordant U-Pb analyses on zircon with upper intercept ages of c. 1.84 and 1.8 Ga. Monazites from both samples gave ages of c. 1.8 Ga (A947 Trekobb, A948 Dyneskär). It is conceivable that these anatectic aplogranites crystallized at c. 1.8 Ga, and some zircon may be inherited from their source rocks. The Rb-Sr muscovite ages for discordant coarse-grained pegmatites also give ages of c. 1.8 Ga. The pegmatite (8 in Table 1) in Västerviken may well be correlated temporally with the 1.8 Ga pegmatites in the Tammisaari area (Hopgood et al. 1983), as well as with pegmatites in Kemiö in southern Finland (Lindroos et al. 1996). Mafic diabase dykes that are associated with pegmatitic granite dykes also represent a late magmatism in the area (Pajunen et al. 2008); it relates to a deep source of the heat flow continuing into the latest Svecofennian evolution. Alviola et al. (2001) described c. 1.79 Ga pegmatitic granite pulses from the Bothnian area in western Finland.

Stabilization of the crust in the south took place shortly after 1.8–1.79 Ga (e.g. Suominen 1991), which represents the age of post-tectonic granites in the southern Svecofennian terrane.

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REFERENCES


